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ORIGINAL PAPER

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Electron-microprobe dating of monazites from Western Carpathian basement granitoids: plutonic evidence for an important Permian rifting event subsequent to Variscan crustal anatexis

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Abstract Accessory monazites from 35 granitoid samples from the Western Carpathian basement have been analysed with the electron microprobe in an attempt to broadly constrain their formation ages, on the basis of their Th, U and Pb contents. The sample set includes representative granite types from the Tatric, Veporic and Gemeric tectonic units. In most cases Lower Carboniferous (Variscan) ages have been obtained. However, a much younger mid-Permian age has been recorded for the specialised S-type granites of the Gemeric Unit, and several small A- and S-type granite bodies in the Veporic Unit and the southern Tatric Unit. This distinct Permian plutonic activity in the southern part of the Western Carpathians is an important, although previously little considered geological feature. It appears to be not related to the Variscan orogeny and is interpreted here to reflect the onset of the Alpine orogenic cycle, with magma generation in response to continental rifting. The voluminous Carboniferous granitoid bodies in the Tatric and Veporic units comprise S- and I-type variants which document crustal anatexis accompanying the collapse of a compressional Variscan orogen sector. The Variscan magmas were most likely produced through the remelting of a subducted Precambrian volcanic arc-type crust which included both igneous and sedimentary reworked volcanic-arc material. Although the 2σ errors of the applied dating method are quite large and typically ±10–20 Ma for single samples, it would appear from the data that the Variscan S-type granitoids (333–367 Ma) are systematically older than the Variscan I-type granitoids (308–345 Ma). This feature is interpreted in terms of a prograde temperature evolution in the deeper parts of the post-collisional Variscan crust. In accordance with recently published zircon ages, this study shows that the Western Carpathian basement must be viewed as a distinct “eastern” tectonomagmatic province in the Variscan collision zone, where the post-collisional crustal melting processes occurred ~20 Ma earlier than in the central sector (South Bohemian Batholith, Hohe Tauern Batholith).

Keywords Western Carpathians · Monazite · Geochronology · Granitoids · Variscan orogeny · Permian rifting

Introduction

The ability of the present generation of electron microprobes to analyse Pb at the 0.1-wt% level accurately has caused a renewed interest in the chemical age dating of U- and Th-rich minerals. In particular, the abundant accessory mineral monazite has turned out to be extremely suitable for this dating technique due to its mostly undisturbed U–Th–Pb system, with typically high radiogenic lead contents and subordinate portions of common lead (cf. Parrish 1990; Suzuki et al. 1991; Montel et al. 1996).

Although chemical dating of monazite with the electron microprobe (EMP) can, of course, not compete in terms of precision and reliability with modern monazite or zircon ages obtained by isotope dilution/mass spectrometry or SHRIMP, its application can be useful in areas where little previous geochronological work has been undertaken, in order to obtain a first quick overview. Here we report EMP monazite ages for basement granitoids from the Western Carpathians. These granitoids have been investigated in recent years in some detail with reference to their petrology, chemistry, possible sources and tectonothermal environments (Petrík et al. 1994; Petrík and Kohút 1997; Petrík 2000 and references therein). A correlation of the petrogenetic data with a geochronological time scale is important in order to un-

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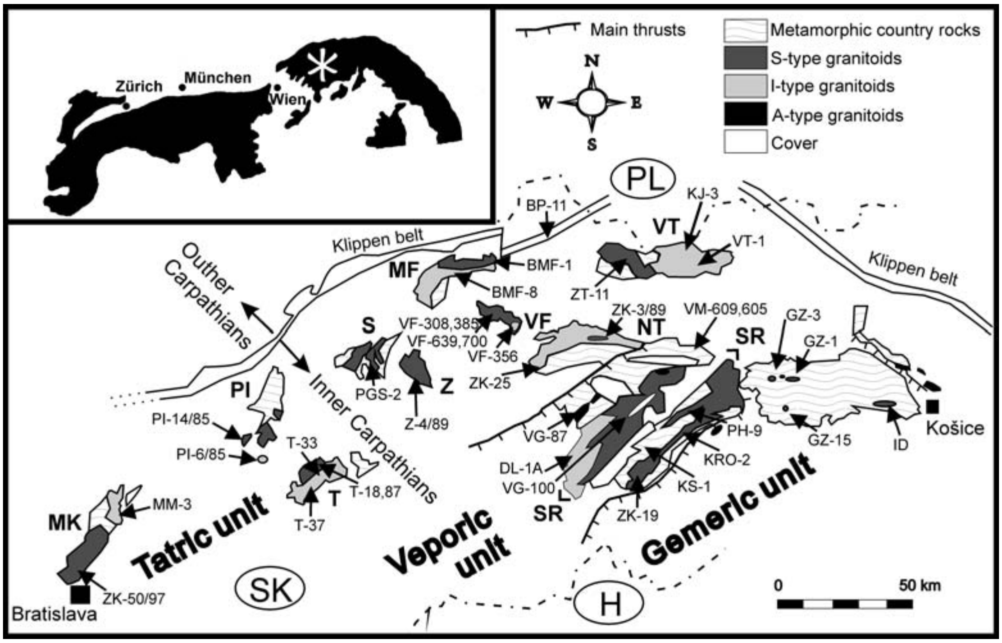

Fig. 1 Distribution of granitoid rocks in the Western Carpathian basement and sample locations of the present study. Abbreviations of basement massifs ("the core mountains"): *MK* Malé Karpaty Mts., *PI* Považský Inovec Mts., *T* Tríbeč Mts., *S* Suchý Massif, *Z* Ziar Mts., *MF* Malá Fatra Mts., *VF* Velká Fatra Mts., *NT* Nízke Tatry Mts., *VT* Vysoké Tatry Mts., *SR* Slovenské Rudohorie Mts. *Inset* shows the position of the Western Carpathians (*star*) within the Alpine-Carpathian orogen

derstand the processes involved in the basement evolution of the Western Carpathians.

Geological background

The Western Carpathian section of the Alpine-Carpathian arc is commonly divided into three major, Alpine basement-bearing tectonic units which were juxtaposed through north-directed thrusting during the Upper Cretaceous. These are, from north to south, the Tatric, Veporic and Gemeric units (Fig. 1). Traditionally, the Western Carpathian basement is considered in a broad sense to be "Variscan". Basement granitoids are most abundant in the Tatric and Veporic units. They are exposed in horst structures which were exhumed during the Alpine orogeny in the Eocene-Miocene, and they are termed the "core mountains" (Fig. 1). Mostly an older roof of metamorphic material (orthogneisses, paragneisses and amphibolites) is preserved around the granitoid bodies.

The effects of Alpine regional metamorphism were relatively minor in the Tatric Unit at very low- to low-grade PT conditions but reached amphibolite facies grades in the Veporic Unit (Plašienka et al. 1999; Janak et al. 2001). The latter constitutes a certain danger for the method of monazite dating because of the potential metamorphic disturbance of the U–Th–Pb system. The basement of the Gemeric Unit consists mainly of low-grade metamorphic formations with only small granite bodies. Both experienced Alpine greenschist facies regional metamorphism (Faryad 1992).

The post-Variscan history of the Western Carpathians involved continental rifting in the Permian and the opening of the Meliata Ocean in the Triassic (Plašienka et al. 1997). Many geologists regard this as the onset of the Alpine orogenic cycle (e.g. Neubauer et al. 2000). However, there are other authors who consider this phase of Permo-Triassic extension as a direct consequence of late Variscan plate tectonics, caused by back-arc spreading behind a northward-dipping subduction system at the southern Variscan fold belt flank (e.g. Stampfli et al. 2001).

Lithology of the basement granitoids

Following the classification of Chappell and White (1974), the Western Carphathian basement granitoids have been divided into S-types and I-types (Cambel and Petrík 1982; Petrík et al. 1994; Petrík and Kohút 1997; Broska and Uher 2001). Most "core mountains" contain both S- and I-type granite units (Fig. 1).

The S-type units comprise mainly granites to leucogranites and a few tonalites to granodiorites, and are characterised by moderately peraluminous compositions. They contain Al-rich biotite and, except for some tonalites, primary muscovite and occasionally sillimanite or garnet. According to the petrological investigations of Petrík and Broska (1994), the S-type melts had low water contents, reduced oxygen activities, and they developed a distinctive accessory mineral paragenesis of apatite, monazite, ilmenite and zircon with dominant (211) + (110) faces.

The I-type units comprise mainly tonalites and granodiorites with metaluminous to weakly peraluminous (subaluminous) compositions. Mafic minerals are Al-poor biotite ± hornblende. The accessory mineral paragenesis is typically zircon + allanite + magnetite + sphene + epidote. Monazite is comparably rare. However, as shown in this study, small monazite grains can mostly be found as well. Higher oxygen activities and water contents have been inferred for the I-type melts by

Petrík and Broska (1994). However, in several cases a clear distinction between S- and I-type granitoids is not possible, and continuous gradations seem to exist between both groups (Cambel and Petrík 1982).

Sr and Nd isotope data (Petrík and Kohút 1997; Kohút et al. 1999; Poller et al. 1999a; Petrík 2000; Poller et al. 2001), together with the widespread presence of inherited zircon components (e.g. Broska et al. 1990; Michalko et al. 1998; Poller et al. 1999b), suggest that both the S-type and I-type granitoids were derived mainly from crustal sources. In the εNd–εSr diagram, fields for the S- and I-type granitoids cover almost the same range, with Sr_i from ~0.705 to 0.710, and εNd_i from –1 to –5. Similarly to the I-types, the S-types often display high Sr and low Rb concentrations which point to little evolved crustal sources, probably volcanic arc-type crust. An exception are the S-type granites of the Gemeric Unit which have significantly higher Rb/Sr and Sr initial ratios (0.710–0.713) and were probably derived from a more evolved, metasedimentary crustal source (Petrík and Kohút 1997; Kohút et al. 1999; Petrík 2000).

Furthermore, a distinct group of small plutons broadly matching the A-type granite classification (Collins et al. 1982; Eby 1990) has been recognised in the Western Carpathians. These have been found in the Veporic and Gemeric units but not in the Tatric basement, and have been considered to be post-orogenic Variscan plutons (Uher and Broska 1996). Boulders of such A-type granites have also been identified in the Cretaceous flysch of the Klippen belt. The A-type melts were reduced, slightly peraluminous and high in fluorine (Uher and Broska 1996).

Not included in our study are the granitoid orthogneisses which occur in the metamorphic roof formations of the granitoids, because of their presumably complex monazite systematics and the difficulty to distinguish between metamorphic and magmatic monazite ages. According to Petrík and Kohút (1997), these rocks are largely of the S-type.

Previous geochronological data

Geochronological work on the Western Carpathian granitoids commenced in the late 1950s with the K–Ar method (Kantor 1959), followed by Rb–Sr mineral and whole-rock data from the late 1960s onwards. Based on this early work, it has been proposed that the basement granitoids are between ca. 250 and 400 Ma old (Cambel et al. 1990; Kohút et al. 1996 and references therein). However, Král (1994) noted that, due to an incomplete homogenisation of the Rb–Sr system during melting and magma mixing effects, the Rb–Sr WR isochrons from Carpathian granitoids are probably too old in many cases, and not capable of providing reliable and precise formation ages (see also discussion in Petrík 2000).

U–Pb zircon ages, which have been obtained from a few localities over the past 12 years, also suggest that the Western Carpathian granitoids formed between the Devonian and Permian. It would appear from these data that the S-type granitoids formed close to 350 Ma (Shcherbak et al. 1990; Michalko et al. 1998) whereas for the I-type granitoids much younger formation ages, close to 300 Ma, have been suggested (Bibikova et al. 1988, 1990; Broska et al. 1990). However, these ages for the I-type granitoids were often inferred from the ^{206}Pb/^{238}U ratios of data points considered as concordant but having large errors in the ^{207}Pb/^{235}U ratios. Clearly, these data bear a high uncertainty. Likewise, the geological significance of some lower intersect ages (e.g. Michalko et al. 1998) remains unclear. Recently, CL-controlled single-grain zircon dating has provided ages between ~360 and ~340 for S-type granites from the Western Tatric Mts., and ages of ~335 and ~315 Ma for I-type granitoids from the High Tatra (Poller et al. 1999b, 2000a). S-type granitoids from Velká Fatra gave a concordant monazite age of 340±2 Ma and a zircon age of 337±11 Ma (Kohút et al. 1997).

For one of the A-type plutons (Hronćok granite), zircon geochronology has provided both a mid-Permian, upper intersect age (Kotov et al. 1996) and a Triassic, lower intersect age (Putiš et al. 2000). Zircons from an A-type granite boulder in the Klippen belt provided a well-defined five-point discordia with an upper intersect age of 274±13 Ma (Uher and Puskharev 1994). All these data suggest that the A-type granites in the Western Carpathians are comparably young. Only the Gemeric S-type granites have been considered to be of a similar Permian age, based on Rb–Sr WR data (Cambel et al. 1989). For these granites, however, a Cretaceous intrusion age has also been considered possible (Vozár et al. 1996; Vozárova et al. 2000). Recently obtained single-grain zircon ages confirm the Permian age estimate (Poller et al. 2000c).

For orthogneisses from the Tatra Mts., protolith ages of ~360–400 Ma have been determined (Poller et al. 2000a). The age of Variscan regional metamorphism in the Tatra Mts. has been constrained at 356±7 Ma by concordant zircon data from migmatitic paragneisses (Poller and Todt 2000).

The sample set for the present study

Figure 1 illustrates where the samples for the present study have been taken. The sample locations and brief rock descriptions are given in Table 1. Representative granite types from the different “core mountains” were selected, so that a systematic comparison of I-type and S-type subunits could be made. Although the I-type granitoids contain on average less and smaller monazite than the S-types granitoids, only a few of the major I-type bodies could not be dated because no monazite was found (e.g. the Shila tonalite which is widespread in the Slovenské Rudohorie Mts., or the Dumbier granite from the Nízke Tatry Mts.).

In the Tatric Unit, representative I- and S-type granitoids were collected from the Malé Karpaty Mts., Tríbeč

Table 1 The dated samples: locations (cf. Fig. 1) and brief petrographic characteristics (references: *1* Macek et al. 1982; *2* Broska and Uher 1988; *3* Broska et al. 2000; *4* Broska and Gregor 1992; *5* Broska et al. 1997; *6* Kohút 1992; *7* Uher et al. 1994; *8* Petrík et al. 1995; *9* Hraško et al. 1997; *10* Finger and Broska 1999)

Sample	Location	Brief characteristics	Type	Ref.
Tatric Unit				
ZK-50/97	Malé Karpaty Mts., Železná Studnička quarry	Ms–Bt granodiorite ("Bratislava granite")	S	1
MM-3	Malé Karpaty Mts., Harmónia, small, abandoned quarry	Bt granodiorite ("Modra granite")	I	
PI-14/85	Považský Inovec Mts., Moravany nad Váhom, Striebornica valley	Ms–Bt granodiorite	S	2
PI-6/85	Považský Inovec Mts., Hlohovec, Stará Hora, road-cut	Leucotonalite	I	2
T-18	Tríbeč Mts., Dršna valley, 2,500 m SE from Krnča village	Bt granodiorite	S	3
T-87	Tríbeč Mts., 1,500 m SE from Krnča village, forest road-cut	Bt tonalite	S	3
T-37	Tríbeč Mts., Velčice 2,800/2,000 m from elevation point Malá Kurňa	Leucogranite	I	4
T-33	Tríbeč Mts., elevation point Javorový hill	Ms granite injections into mylonites	S	
PGS-2	Suchý Mts., upper end of the Liešťany valley	Pegmatoid leucogranite	S	
Z-4/89	Ziar Mts., abandoned Brezany quarry	Bt granite	S	
BMF-1	Malá Fatra Mts., Bystrička quarry	Bt granodiorite	S	5
BMF-8	Malá Fatra Mts., Lipovec, Hoskora valley	Ms–Bt granite	I	5
VF-308	Velká Fatra Mts., Stanova valley, forest road-cut	Bt granodiorite ("Kornietov type")	S	6
VF-639	Velká Fatra Mts., Blatná valley, natural outcrop	Bt granodiorite ("Kornietov type")	S	6
VF-385	Velká Fatra Mts., Nižná Lipová, cliff on ridge	Ms–Bt granite ("Lipová type")	S	6
VF-700	Velká Fatra Mts., Lower Matejkovo valley, prospecting gallery	Ms leucogranite ("Lubochňa type")	S	6
VF-356	Velká Fatra Mts., Upper Matejkovo valley, abandoned quarry	Bt tonalite ("Smrekovica type")	I	6
ZK-3/89	Nízke Tatry Mts., 100 m W of terminus of the funicular to Chopok hill	Bt granite	S	1
ZK-25	Nízke Tatry Mts., road Sopotnica-Hronov, end of Sopotnica valley	Bt granodiorite	I	1
ZT-11	Western Tatra, cliff next to the highest Rohače lake	Leucogranodiorite	S	
VT-1	High Tatra Mts., Štrbské Pleso, cliff of waterfall "Skok"	Bt tonalite	I	
KJ-3	High Tatra Mts., Dolina Rybiego, Potoku valley, Poland	Leucotonalite	I	
BP-11	Klippen belt, Široká, 1,700/2,840 m from elevation point Turfkov Ziar (851 m)	Granite boulder	S	7
Veporic Unit				
PH-9	Slov. Rud. Mts., 5 km SW Muráň, 850 m NW elevation point 1,018	Bt granite	S	
ZK-19	Slov. Rud. Mts, Road Poltár-České Brezovo, first quarry on east side	Ms–Bt granite ("Rimavica type")	S	1
KRO-2	Slov. Rud. Mts., Krokava, road cut 500 m below the chalet	Leucogranite vein in Rimavica granite	S	
VG-100	Slov. Rud. Mts., Ráztočno, cliff 1 km S of Klenovský Vepor hill	Leucogranodiorite	S	
DL-1A	Slov. Rud. Mts., 7 km NNE Hriňová, small quarry in Slatina valley	Leucogranite dike in Sihla tonalite	I	
VG-87	Slov. Rud. Mts., Kamenistá valley, Hrončok gamekeeper	Mylonitised leucogranite ("Hrončok type")	A	8
KS-1	Slov. Rud. Mts., Klenovec village, borehole KS-1, depth 528–531 m	Ms–Bt granite ("Klenovec type")	S	9
VM-605	Nízke Tatry Mts., main ridge, 1 km W from the Kráľova Hoľa	Strongly sheared leucogranite	S	
VM-609	Nízke Tatry Mts., southern slope of Orlová saddle, 1,750 m a.s.l.	Coarse-grained mylonitised granite	S	
Gemeric Unit				
GZ-1	Slovenské Rudohorie, Hnilec cliff, 800 m from Peklisko elevation point	Leucogranite	S	10
GZ-3	Slovenské Rudohorie, Hnilec, 220 m NE from elevation point Surovec	Leucogranite	S	10
GZ-15	Slov. Rud. Mts., Betliar, cliff 3,250 m SW from elevation point Volovec	Leucogranite	S	10
ID	Road Košice-Zlata Idka, 13 km W Košice, borehole ID-2, depth 132 m	Bt-tourmaline granite	S	

Mts., Považský Inovec Mts., Malá Fatra Mts., Velká Fatra Mts., Vysoké Tatry (High Tatra) Mts., and the Tatric part of the Nízke Tatry (Low Tatra) Mts. Samples of S-type granites were taken from the Ziar and Suchy massifs, and an S-type granitic boulder from the Klippen belt was sampled as well (BP-11).

In the Veporic Unit we investigated the A-type Hronćok granite (VG-87), a leucocratic dike (DL-1A) considered as related to the major I-type unit in the Slovenské Rudohorie Mts. (Shila granite), and representative samples from the main S-type granite units of this area (ZK-19, ZK-100). Furthermore, small S-type granite bodies and dikes from the Slovenské Rudohorie Mts. were sampled (PH-9, KRO-2, KS-1). From the Kráľova Hoľa Massif (Veporic part of the Nízke Tatry Mts.), two S-type granite samples were collected (VM-605 and VM-609).

In the Gemeric Unit four granite samples from the Hlinec, Betliar and Zlata Idka stocks were examined.

Results

General aspects

Between 3 and 18 monazite analyses were carried out per sample. All obtained Th, U and Pb concentrations are given in the Appendix, together with the calculated model ages and 2σ errors. It can be seen from this compilation that for single samples, the monazite model ages of all analysis points are mostly consistent with each other and overlap within their 2σ errors. There were only two exceptions. In sample T-33 one monazite showed significantly older model ages in its centre. This obviously inherited core has been excluded from the average age calculation. In sample DL-1, single monazite analyses give unrealistically low ages, deviating by more than 2σ from the mean value. We attribute this to local lead loss or Alpine age recrystallisation/over-growth effects and omitted these in the average calculations as well. Since isotope ratios cannot be determined, however, we are aware that minor effects of inheritance, common lead presence or lead loss on the mean age can not be ruled out with certainty, and one has to rely on the (fortunately well-established) empirical rule that, in the case of the mineral monazite, these effects are usually not large.

Furthermore, it should be mentioned that in the Veporic granites the monazites often showed marginal alterations with the growth of secondary apatite and allanite, as a consequence of the Alpine reheating (Broska and Siman 1998). For dating, only the largest and best preserved monazite relics were used in such cases, and analyses points were placed at least 10 µm away from the alterations.

The average ages for all dated samples are compiled in Table 2. For six granitoids of this sample set, zircons ages became recently available as well (see annotations in Table 2). The chemical monazite ages match with these zircon ages in all cases. This suggests that the obtained monazite ages can be generally taken as reliable and geologically meaningful.

The S-type granitoids

The data show that at least two generations of S-type plutons are present in the Western Carpathian basement – Permian and Lower Carboniferous ones. Permian ages were obtained for all four investigated samples of granites from the Gemeric Unit. Additionally, the S-type granites VM-609 from the Nízke Tatry and KS-1 (Klenovec granite) from the Slovenské Rudohorie Mts. (both Veporic Unit) gave a Permian age. Finally, a similarly young age (273±17 Ma) was obtained for a distinct, small S-type granite occurrence in the Tríbeč Massif (Tatric Unit), which was already considered as relatively younger on geological grounds because it intruded late Variscan mylonites.

All other investigated S-type granites from the Veporic and Tatric units provided much higher chemical monazite ages. The highest values were obtained from S-type granites from the Považský Inovec Massif (364±17 Ma) and the Slovenské Rudohorie Mts. (369±30 Ma), whereas the lowest (in this Lower Carboniferous group) was obtained for the Kornietov granodiorite from the Velká Fatra Mts. (333±24 Ma). However, the relatively high errors inherent to the method of EMP monazite dating do not allow it to be resolved whether the Lower Carboniferous S-types intruded in two or more independent pulses between ~360 and 330 Ma, or all very close to ~350 Ma. In any case, the monazite ages confirm the previous view, which was mainly based on Rb–Sr whole-rock dating and few zircon data, that in the Western Carpathians an important phase of S-type granite formation occurred at the beginning of the Carboniferous.

Table 2 Average monazite ages (errors at 95% C.L.) for single samples considered as dating granite formation

Location	Sample	Type	Age (Ma)
Tatric Unit			
Malé Karpaty Mts. (MK)	ZK-50/97	S	355±18
	MM-3	I	345±22
Považský Inovec Mts. (PI)	PI-14/85	S	364±17
	PI-6/85	I	323±22
Tríbeč Mts. (T)	T-18	S	357±13
	T-87	S	352±17
	T-37	I	331±22
	T-33	S	273±17
Suchý Massif (S)	PGS-2	S	342±13
Ziar Mts. (Z)	Z-4/89	S	348±22
Malá Fatra Mts. (MF)	BMF-1	S	342±18
	BMF-8	I	336±9
Velká Fatra Mts. (VF)	VF-308	S	333±24
	VF-639	S	348±21
	VF-385	S	348±18
	VF-700	S	343±12
	VF-356	I	308±30[a]
Nízke Tatry Mts. (NT)	ZK-3/89	S	362±27
	ZK-25	I	326±31
Vysoké Tatry Mts. (VT)	ZT-11	S	347±24[b]
	VT-1	I	327±28[c]
	KJ-3	I	317±15[d]
Klippen Belt	BP-11	S	348±22
Veporic Unit			
Slov. Rudohorie Mts. (SR)	PH-9	S	357±21
	ZK-19	S	352±13
	KRO-2	S	367±34
	VG-100	S	369±30
	DL-1A	I	321±18
	VG-87	A	263±19[e]
	KS-1	S	266±16
Nízke Tatry Mts. (NT)	VM-605	S	359±17
	VM-609	S	269±22
Gemeric Unit			
Slov. Rudohorie Mts. (SR)	GZ-1	S	272±11
	GZ-3	S	276±13
	GZ-15	S	273±13
Eastern part	ID	S	263±28[f]

Annotations a–f refer to zircon ages obtained from the same rock type

[a] 304±2 Ma (Poller et al. 2000b)
[b] 363±11 Ma (Poller et al. 1999b)
[c] 332±15 Ma (Poller et al. 1999b)
[d] 315±5 Ma (Poller et al. 1999b)
[e] 278±11 Ma (Kotov et al. 1996)
[f] 265±20 Ma (Poller et al. 2000c)

The I-type granitoids

The investigated I-type granitoid samples provided chemical monazite ages between 345±22 Ma (Modra Massif, Malé Karpaty) and 308±30 Ma (tonalite Velká Fatra Mts.). However, these values are not in agreement with the earlier concepts based on zircon dating, according to which most I-type granitoids in the Western Carpathians formed at ~300 Ma (see compilation in Petrík and Kohút 1997). Some of the I-types, e.g. samples MM-3, T-37 and ZK-25, provided monazite ages which, within their errors, would be even compatible

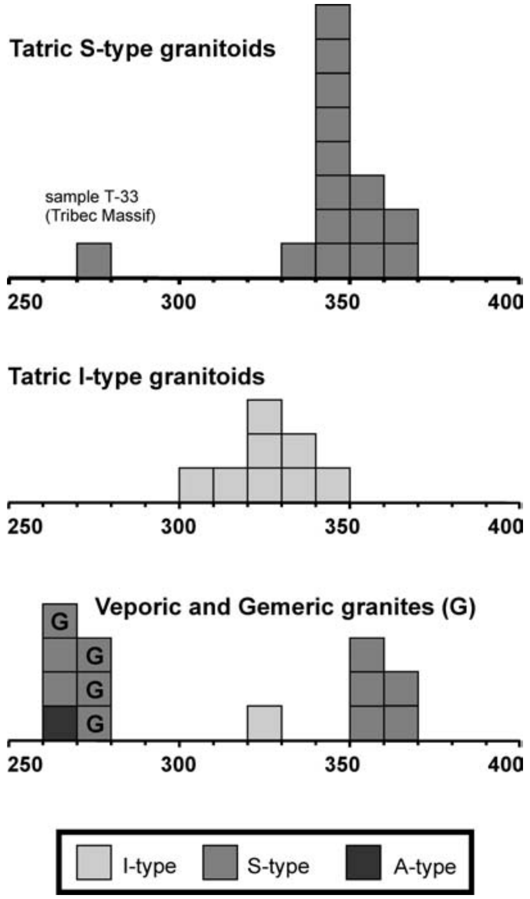

Fig. 2 Histogram showing the distribution of EMP monazite ages obtained from S-, I- and A-type granitoid occurrences in the Tatric, Veporic and Gemeric units (data source, see Table 2)

with a formation around 350 Ma, coeval with the S-types. Other samples, such as BMF-8, PI-6/95 and DL-1A, provide ages clearly younger than 350 Ma, pointing to a late Lower Carboniferous or early Upper Carboniferous plutonic activity. From the Polish part of the High Tatra, Finger et al. (2000) have recently reported a chemical monazite age of 317±15 Ma for a leucotonalite sample (KJ-3 in Table 2). The graphic compilation of ages in Fig. 2 suggests that I-type granite formation generally postdates the early Carboniferous phase of S-type granite formation.

The Hrončok A-type granite

For this rock we obtained an age of 263±19 Ma. This is close to the zircon age of 278±11 Ma given by Kotov et al. (1996) for a subvolcanic dyke derived from this granite (Petrík 1996), although it is slightly higher than the Triassic zircon age (239±1 Ma) of Putiš et al. (2000) for a sample from the main Hrončok granite body. Nevertheless, it is clear also from our data that the Hrončok granite complex must be post-Carboniferous.

Discussion and conclusions

Variscan granite formation in the Western Carpathians

There is now good geochronological evidence that many of the Variscan granitoids of the Western Carpathians formed in the Lower Carboniferous through voluminous melting of crust during, or soon after, collision-related Variscan crustal thickening and regional metamorphism. From the Tatra Mts., Janak et al. (1999) have reported clear evidence for decompression melting of paragneisses during their post-collisional uplift, producing syndeformational migmatites with an age of 356±7 Ma (Poller and Todt 2000). This shows that the crust was in a partially molten state at that time. Sr and Nd isotope data (Kohút et al. 1999; Poller et al. 1999a, 2001) rule out that these metapelites/metagreywackes were the main sources of the Tatric and Veporic plutons. However, the same process of decompression melting may have affected less evolved sources in deeper parts of the crust, with melts segregating and rising into higher crustal levels.

It is likely that temperatures in the middle and lower levels of the uplifting orogen further increased in the following few million years due to radiogenic heat production (e.g. Gerdes et al. 1999) and a heat input from the mantle, which very often accompanies the collapse stage of collision-type orogenies (e.g. Henk et al. 2000). More precise dating methods will now be needed to resolve the exact timing of the Carboniferous plutonic activity in the Western Carpathians. From our data set it would appear that the age difference between S- and I-type granitoids is significant but perhaps not as great as previously thought. Zircon ages around 300 Ma, calculated for I-type granitoids from the Tríbeč and Sihla massifs from ^{238}U/^{206}Pb ratios, may be too young due to lead loss effects. Taking into account the recent 315- and 335-Ma U–Pb zircon ages of Poller et al. (1999b) for I-types in the High Tatra, there is some evidence for major, I-type granite-forming event(s) in the Visean and early Upper Carboniferous. Indeed, post-collisional I-type plutons postdating regional metamorphism for some 20–40 Ma are common in collisional orogens world-wide (Pitcher 1983; Harris et al. 1984).

Both S- and I-type granitoids in the Western Carpathians often have chemical features which indicate remelting of volcanic arc-type crust (low Rb, high Sr, low Y and HREEs, and deep negative Nb anomalies relative to Ce and Th). Weathered and sedimentary reworked Precambrian/Cadomian volcanic-arc material may have been the source for many of the Tatric and Veporic S-type granitoids. Such mica-enriched portions of subducted arc-type crust would be the first to melt in a (post-)collisional high heat flow regime due to fluid-absent melting reactions of the type Mu + Qz + Plag → melt + Sil ± Kfsp or Bt + Sil + Qz + Plag → melt ± Kfsp ± Grt/Crd (see Clemens and Vielzeuf 1987).

On the other hand, the unweathered, less aluminous portions of the same volcanic-arc crust (the I-type sources) may have remained solid, and then melted a couple of million years later due to a further temperature rise involving reactions such as Bt + Plag + Qz → melt ± Grt ± Opx ± Kfsp, or hornblende dehydration melting. Such a model could well explain the observed time differences between S- and I-type plutons. Furthermore, the early-stage melts may have received some (minor) admixtures from country-rock paragneiss leucosomes at their emplacement level, which may have shifted their composition even more towards the S-type.

Intra-Variscan correlations

Although the chain of events, i.e. high/medium-P burial regional metamorphism → exhumation → decompressional anatexis and production of crustal granites → intrusion of post-collisional I-type plutons, appears to be basically the same as in other major Variscan granite terrains in central Europe, it should be noted that in the Western Carpathians the whole process started relatively early. The time scale of tectonothermal events recorded in the Carpathians is consistently shifted for some 20–30 Ma towards older ages compared to the westerly adjacent Variscides, i.e. the Bohemian Massif in Austria and the Czech Republic (Fig. 3). There, regional metamorphism occurred at ca. 340 Ma, followed by rapid uplift and extensive crustal melting between about 335 and 320 Ma (formation of most of the South Bohemian Batholith; Finger et al. 1997). A later phase of I-type plutonism is dated at ca. 315–300 Ma, with some late I-type dyke swarms being as young as 270 Ma (Košler et al. 2001).

The Variscan granite terrain of the Western Carpathians is also distinct from the Variscan Hohe Tauern Batholith in the Eastern Alps (Finger et al. 1993) with regard to the age of magmatic pulses (Fig. 3). Large-scale crustal melting in the Hohe Tauern occurred at ca. 330–340 Ma (Eichhorn et al. 2000). After some 30 Ma of magmatic quiescence, a new, intensive pulse of I-type plutonism is recorded at around 290–310 Ma (Cesare et al. 2001).

This unconformity of post-collisional magmatic events indirectly supports the palinspastic reconstructions of Stampfli (1996) and Von Raumer (1998), according to which, in the Visean, the Western Carpathian rocks were positioned far away, i.e. some 500 km east of the Bohemian Massif and the Hohe Tauern. Evidently, the tectonic history of this "east sector" of the Variscan fold belt was quite distinct. We may speculate that the Western Carpathian basement belonged to those Variscan (Armorican) terranes which collided first with the Laurasian megacontinent. It remains open for discussion whether a correlation is feasible with those extra-Alpine Armorican terranes, which docked relatively early to Laurasia as well (e.g. the Saxothuringian and the Teplá-Barrandian). Indeed, there are certain parallels in the Variscan evolution of the Carpathian terrane and the Teplá Barrandian Unit of western Bohemia (the Bohemian terrane of Franke and Zelazniewicz 2000). There, collisional metamorphism occurred at roughly 370 Ma, followed by I-type plutonism at ca. 350 Ma (Dörr et al. 1998). However, it should be noted that, unlike the Western Carpathians, the Bohemian terrane is very poor in S-type granitoids.

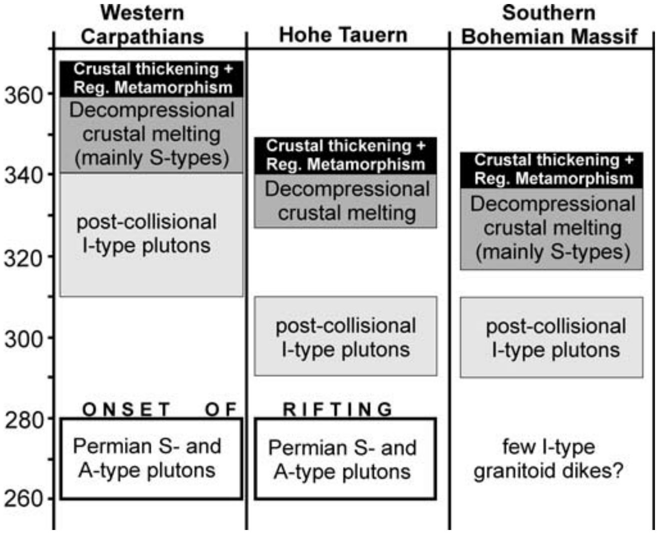

Fig. 3 Timing of Variscan tectonomagmatic events in the Western Carpathians compared to the Hohe Tauern granitoid terrain and the Southern Bohemian Massif (Moldanubian; see text for data sources)

A broadly corresponding timing of Variscan tectonothermal events may have existed in the Western Carpathians and the westwards adjacent Lower and Middle Austroalpine units of the Eastern Alps. The latter contain Variscan I-type and S-type plutons with compositions similar to the Carpathian ones (see data in Schermaier et al. 1997). Unfortunately, reliable geochronological information is presently scarce for these Austroalpine plutons. From Rb–Sr WR data (Scharbert 1990; Peindl et al. 1990), it would seem that the largest Austroalpine S-type pluton, which is represented by the Grobgneis of the Semmering area, is as old as the Western Carpathian S-type granitoids. Likewise, unpublished zircon ages of Von Quadt (personal communication) constrain a phase of post-collisional I-type granite formation in the Middle Austroalpine at about 335 Ma (Seckau-Bösenstein Batholith), which falls in the time span of I-type granite formation in the Western Carpathians. A correlation of the Tatric and Veporic units with the Lower and Middle Austroalpine basement units has been suggested by Neubauer (1994). Together they may indeed constitute one coherent "east sector" in the Variscan collision zone (see Von Raumer 1998).

Tectonic significance of the Permian granites

In the extra-Alpine Variscan massifs (Massif Central, Schwarzwald, Bohemian Massif), Permian granites are very rare whereas recent work, including this paper, has shown that granites of this age are almost ubiquitous in the Alpine-Carpathian chain (e.g. Von Quadt et al. 1999; Schaltegger and Gebauer 1999; Thöni 1999; Eichhorn et al. 2000). Therefore, these Permian granites obviously indicate a new tectonomagmatic event in the intra-Alpine Variscan units, and are not related to the collapse of the Variscan orogen. Due to the scarcity of geochronological data, this has remained unrecognised in a couple of previous studies (e.g. Finger et al. 1997).

The geochemical characteristics of the intra-Alpine (intra-Carpathian) Permian plutonism are extremely variable, indicating the involvement of several different magma sources. Stocks and dykes of leucocratic Permian S- and A-type granites, as present in the southern part of the Western Carpathians, have also been found in the Hohe Tauern and the western Alps (Von Quadt et al. 1999; Schaltegger and Gebauer 1999; Eichhorn et al. 2000 and references therein). At least the S-type granites clearly show that crustal melting has occurred at that time. For the A-type melts different genetic models can be discussed. They may have formed by high-T melting of crust (Collins et al. 1982), or by fractionation of enriched mantle melts (see Bonin 1992).

In particular in the southern and western Alps, several ca. 270-Ma-old I-type plutons (e.g. Dora Maira, Monte Rose, Grand Paradiso) have been recognised (Bussy and Cadoppi 1996; Bertrand et al. 2000). Sources may have been partly crustal (remelting of metaigneous crust) or may contain a mantle contribution (see Schaltegger and Gebauer 1999 and references therein). Furthermore, it should not be overlooked that a considerable part of the Permian magmatism in the Alps is basaltic to gabbroic, with MORB-like or WPB-like composition (e.g. Pin and Sills 1986; Hermann et al. 1997; Miller and Thöni 1997). At the same time, low-pressure metamorphism has been locally documented in metapelitic lithologies (Schuster and Thöni 1996).

Due to the observed style of the magmatic/metamorphic record, most geologists presently agree that the "Permian event" in the Alpine-Carpathian chain is caused by extensional tectonics, involving high heat flow from the mantle through basaltic underplating (Bussy et al. 2000; Broska and Uher 2001; Thöni 1999). Whether this Permian extension was caused by a late Variscan, northwards subduction of the Palaeotethys Ocean (back-arc extension model of Stampfli et al. 2001), or whether we are dealing with an intracontinental rift and the beginning of a new Wilson cycle (onset of the Alpine orogeny in the sense of, e.g. Neubauer et al. 2000) is a matter of debate. In the Italian and Swiss sectors of the Alps, there is some evidence that (a potentially subduction-related) I-type plutonism persisted from the late Carboniferous (Cesare et al. 2001) to the Permian, and one may argue that this corroborates the Stampfli et al. (2001) subduction model. In the Carpathians the situation is different: Here, the late Variscan granites, although of the I-type, do not appear to be subduction related, and there seems to be a considerable time gap between the Variscan and the Permian granites.

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Appendix

Analytical techniques

Monazite analyses were carried out between 1995 and 2001 using the Jeol JX 8600 microprobe of the Institute for Geology at Salzburg University. The monazite grains were searched in polished thin sections by backscattered-electron imaging (BSE). For analysis, the operating conditions were 15 kV and 250 nA with a beam diameter of 5 μm. Analysis spots were preferentially placed in the grain centres. In larger grains, analyses were made in different places (see below). For Th, U, Pb Mα spectral lines and counting times of 30 (2×15), 50 (2×10) and 200 s (2×100 s) were chosen for peak and background positions. The 2σ errors per spot were typically 0.05 wt% for Th, 0.03 wt% for U, and 0.012–0.013 wt% for Pb.

Y, La, Ce (Lα_1) Pr, Nd (Lβ_1) and P, Al, Si, Ca (Kα_1) were routinely analysed to provide a reasonable ZAF correction and to control whether the analysis points had an optimal monazite stoichiometry. Furthermore, slight Y and Th interferences on the Pb Mα line had to be empirically corrected (Montel et al. 1996; Scherrer et al. 2000) as well as a Th interference on U Mα. Calibration standards were synthetic ThO_2, UO_2, PbS, apatite, $CeAl_2$ and a REE glass. To independently test the quality of the Th–U–Pb analyses, a monazite age standard, dated by isotope dilution and mass spectrometer analyses with a concordant age of 341±2 Ma (Friedl 1997), was systematically analysed together with the samples. The results of these standard measurements are given below. The recommended age value could be sufficiently reproduced during all analytical sessions.

Calculation of ages

Ages were calculated for each analysis point with the following equation (Montel et al. 1996):

$$\begin{aligned} Pb = {} & \frac{Th}{232} \times \left(e^{\lambda_{232} \times T} - 1\right) \times 208 \\ & + \frac{U}{238} \times 0.9928 \times \left(e^{\lambda_{238} \times T} - 1\right) \times 206 \\ & + \frac{U}{238} \times 0.0072 \times \left(e^{\lambda_{235} \times T} - 1\right) \times 207 \end{aligned}$$

Appropriate two sigma errors were derived by propagating the individual errors for Pb, Th, U through this equation (Table 3). For each sample, a weighted average age was calculated from all obtained monazite analyses using the software of Ludwig (2000). This age has been generally interpreted as the granite formation age.

Table 3 Th, U, Pb compositions (wt%) and model ages of the analysed monazites, including data for laboratory standard F-5

Grain number/ analysis point	Th	U	Pb	Age
Session 9-95				
Standard F-5				
1.1	12.241	0.840	0.222	332±21
1.2	10.848	0.687	0.196	336±24
1.3	6.800	1.001	0.153	341±31
2.1	6.931	0.878	0.148	340±32
2.2	4.773	0.813	0.115	346±42
1.4	11.217	0.771	0.220	359±23
1.5	6.675	1.124	0.161	350±30
1.6	12.418	0.806	0.243	361±21
1.7	11.383	0.651	0.204	339±23
1.8	5.680	1.127	0.138	331±34
2.3	7.643	0.861	0.161	345±30
Sample VT-1				
1.1	3.361	0.011	0.044	292±92
1.2	3.746	0.091	0.049	272±78
2.1	3.401	0.118	0.063	371±83
3.1	3.279	0.023	0.059	392±93
4.1	4.598	0.034	0.074	351±66
4.2	4.078	0.044	0.066	347±74
5.1	3.534	0.013	0.050	310±88
6.1	3.639	0.047	0.047	280±83
Sample VG-87				
1.1	5.479	0.150	0.067	253±53
1.2	5.504	0.148	0.069	258±52
1.3	5.864	0.172	0.076	266±49
2.1	3.972	0.097	0.049	258±73
2.2	5.344	0.165	0.072	275±53
1.4	5.245	0.161	0.073	283±54
1.5	4.156	0.072	0.052	266±71
3.1	4.401	0.047	0.048	235±69
3.2	4.214	0.058	0.051	260±71
Sample Pl-14/85				
1.1	5.266	0.068	0.088	359±57
1.2	4.827	0.681	0.122	387±44
2.1	4.237	0.065	0.083	418±70
2.2	4.674	0.071	0.079	358±64
2.3	4.695	0.708	0.113	362±45
1.3	4.316	0.055	0.069	341±70
1.4	5.153	0.061	0.081	336±59
3.1	5.522	0.065	0.092	360±55
3.2	4.768	0.502	0.101	352±49
3.3	5.569	0.254	0.104	363±49
Sample ZK-50/97				
1.1	4.197	0.130	0.079	380±68
1.2	3.800	0.160	0.073	375±72
2.1	4.177	0.288	0.087	382±61
3.1	4.179	0.083	0.069	347±70
4.1	3.745	0.245	0.070	344±69
5.1	4.589	0.025	0.070	337±67
5.2	4.186	0.050	0.065	332±72
6.1	3.199	0.210	0.053	303±81
7.1	6.068	0.122	0.102	352±48
7.2	9.157	0.238	0.159	359±32
Sample PGS-2				
1.1	5.232	0.734	0.116	340±41
2.1	3.878	1.581	0.134	333±35
3.1	4.459	1.208	0.130	346±37
4.1	3.777	1.016	0.110	350±44
5.1	3.132	0.808	0.088	343±55
1.2	1.102	0.245	0.032	381±165
2.2	4.070	1.442	0.122	314±36
6.1	3.617	0.337	0.083	392±66
5.2	4.377	1.232	0.136	363±37
7.1	3.685	0.870	0.097	334±48
8.1	3.857	0.066	0.061	337±77
Session 4-96				
Standard F-5				
1.1	9.779	0.766	0.203	371±26
1.2	9.134	0.714	0.177	347±27
1.3	6.174	1.227	0.136	300±31
1.4	11.629	0.758	0.216	342±22
1.5	10.013	0.810	0.204	361±25
1.6	9.949	0.829	0.190	336±25
1.7	10.682	1.015	0.216	346±22
Sample ZK-19				
1.1	6.055	0.533	0.131	377±40
1.2	6.847	0.977	0.146	326±31
1.3	6.198	0.934	0.150	363±34
1.4	6.638	0.652	0.140	359±36
2.1	5.356	0.429	0.108	357±46
2.2	5.571	0.158	0.094	347±51
3.1	6.769	0.939	0.155	354±32
3.2	5.090	0.686	0.111	340±43
Sample VF-356				
1.1	2.728	0.332	0.053	312±83
1.2	1.990	0.061	0.033	335±143
2.1	5.289	0.093	0.075	300±56
2.2	4.060	0.083	0.055	284±72
2.3	2.619	0.567	0.064	321±70
2.4	2.713	0.423	0.059	324±77
Session 3-97				
Standard F-5				
1.1	11.285	0.674	0.213	354±23
1.2	6.704	0.924	0.151	348±32
1.3	6.652	2.295	0.207	329±22
1.4	7.759	0.764	0.150	329±31
1.5	7.711	0.772	0.153	336±31
1.6	3.181	0.069	0.052	340±92
1.7	5.120	0.100	0.084	344±58
Sample DL-1A				
1.1	3.672	0.071	0.036	205±81[a]
2.1	6.143	0.145	0.100	336±47
2.2	5.512	0.246	0.092	327±50
3.1	5.138	0.283	0.087	320±52
4.1	4.672	0.222	0.073	305±58
5.1	5.215	0.278	0.090	328±51
5.2	4.792	0.163	0.050	211±59[a]
2.3	6.921	0.131	0.103	312±43
3.2	6.463	0.158	0.099	318±45
Sample MM-3				
2.1	5.049	0.067	0.084	356±59
2.2	3.354	0.095	0.055	338±86
1.1	2.981	0.074	0.052	359±97
2.3	4.204	0.076	0.066	333±70
3.1	4.593	0.081	0.071	328±64
4.1	4.976	0.034	0.076	333±62
5.1	5.356	0.087	0.092	364±56
6.1	5.605	0.079	0.090	343±53
Sample Z-4/89				
1.1	3.063	0.592	0.077	345±63
1.2	3.109	0.548	0.076	349±64
1.3	3.082	0.559	0.079	361±64
2.1	3.170	0.499	0.073	342±66
3.1	3.519	0.429	0.077	349±64
4.1	3.817	0.333	0.074	336±64
5.1	3.853	0.264	0.072	343±67
2.2	4.150	0.156	0.074	355±67
Sample BMF-1				
1.1	3.296	0.589	0.079	342±60
2.1	2.997	0.592	0.074	335±64
2.2	3.339	0.557	0.077	335±61
3.1	3.657	0.471	0.080	345±60
4.1	3.142	0.388	0.070	354±71
5.1	2.096	0.292	0.048	351±103
6.1	2.886	0.235	0.055	334±86
7.1	4.110	0.111	0.063	316±70
8.1	4.024	0.221	0.082	386±66
3.2	4.352	0.801	0.101	327±45
3.3	4.117	0.371	0.083	351±59
4.2	3.586	0.122	0.062	346±79
4.3	3.809	0.125	0.064	339±74
Sample BMF-8				
1.1	4.610	0.153	0.075	330±61
2.1	9.146	0.356	0.154	334±30
2.2	9.954	0.358	0.171	343±28
4.1	9.622	0.386	0.165	339±29
4.2	7.633	0.369	0.131	332±35
4.3	9.817	0.392	0.166	335±28
4.4	9.723	0.352	0.157	324±29
4.5	9.472	0.384	0.165	345±29
4.6	8.793	0.333	0.141	319±32
4.7	8.026	0.335	0.145	356±34
4.8	7.196	0.302	0.129	353±38
4.9	6.908	0.329	0.114	319±39
Session 4-97				
Standard F-5				
1.1	9.523	0.775	0.184	342±26
1.2	10.115	0.754	0.188	335±25
1.3	7.282	0.980	0.162	346±30
1.4	11.848	0.806	0.218	337±22
1.5	10.078	0.758	0.198	352±25
1.6	7.821	0.935	0.160	330±29
Sample VM-609				
1.1	5.955	0.184	0.087	296±48
1.2	3.810	0.067	0.051	285±78
1.3	3.144	0.057	0.042	285±94
1.4	3.880	0.082	0.046	246±76
2.1	4.020	0.070	0.053	277±74
2.2	4.330	0.100	0.057	275±67
3.1	4.837	0.112	0.058	252±60
3.2	5.651	0.180	0.068	244±50
Sample VM-605				
1.1	5.032	0.151	0.085	343±57
1.2	10.560	0.710	0.215	374±24
1.3	10.463	0.776	0.197	340±24
2.1	9.465	0.806	0.200	370±26
2.2	8.824	0.786	0.170	334±28
3.1	7.114	1.109	0.182	379±29
3.2	6.143	0.926	0.148	362±34
Session 9-97				
Standard F-5				
1.1	5.127	1.349	0.145	342±33
1.2	5.340	1.104	0.132	333±35
1.3	5.067	1.392	0.149	349±33
1.4	9.097	0.830	0.173	328±27
1.5	10.645	0.819	0.200	336±24
1.6	9.726	1.140	0.210	350±23
1.7	5.217	1.317	0.145	342±33
Sample T-18				
1.1	8.627	0.258	0.156	368±33
2.1	7.848	0.213	0.124	324±37

Table 3 (continued)

Grain number/ analysis point	Th	U	Pb	Age
3.1	4.519	0.178	0.095	418±61
4.1	4.383	0.186	0.071	317±63
5.1	4.883	0.195	0.071	289±57
6.1	5.471	0.192	0.103	377±51
7.1	6.196	0.229	0.115	370±45
8.1	6.416	0.198	0.125	395±44
1.2	8.673	0.249	0.162	381±33
2.2	12.488	0.292	0.232	386±23
1.3	12.907	0.307	0.216	347±23
2.3	11.378	0.265	0.200	364±26
3.2	10.004	0.258	0.180	371±29
4.2	11.722	0.292	0.186	328±25
5.2	11.034	0.283	0.190	356±26
6.2	9.457	0.266	0.147	319±30
7.2	7.648	0.224	0.133	354±37
8.2	7.136	0.239	0.124	351±40
Sample T-87				
1.1	6.082	0.461	0.125	369±41
1.2	6.858	0.497	0.132	348±37
1.3	6.173	0.442	0.120	353±41
2.1	3.526	0.317	0.072	353±69
2.2	2.837	0.278	0.068	407±84
3.1	7.366	0.489	0.136	340±35
3.2	3.803	0.361	0.071	321±63
Sample T-37				
1.1	4.566	0.252	0.089	371±58
2.1	4.783	0.282	0.078	305±55
3.1	4.068	0.264	0.081	369±64
4.1	4.061	0.246	0.070	325±65
5.1	4.674	0.364	0.082	314±54
6.1	4.438	0.213	0.071	311±61
6.2	4.415	0.327	0.081	332±57
Session 5-98				
Standard F-5				
1.1	12.707	0.971	0.246	348±20
1.2	10.433	0.414	0.184	350±27
1.3	23.412	0.983	0.406	342±12
1.4	0.920	0.234	0.017	226±188
1.5	12.567	0.953	0.242	346±20
1.6	10.252	0.371	0.172	336±27
1.7	23.540	0.955	0.405	340±12
1.8	0.967	0.222	0.032	427±185
1.9	1.093	0.246	0.022	267±167
Sample BP-11				
1.1	4.640	0.135	0.081	355±62
2.1	4.353	1.064	0.118	339±40
2.2	3.429	0.273	0.075	387±72
3.1	4.046	0.183	0.070	336±68
4.1	3.981	0.216	0.072	345±67
5.1	3.866	0.347	0.067	303±63
5.2	3.526	0.325	0.081	396±68
Sample GZ-1				
1.1	10.618	0.762	0.149	256±24
2.1	11.489	1.213	0.192	279±20
3.1	13.340	1.371	0.219	276±18
Sample PI-6/85				
1.1	4.170	0.125	0.077	378±69
2.1	4.914	0.116	0.067	283±59
3.1	5.097	0.148	0.083	335±56
4.1	5.312	0.353	0.084	291±48
4.2	5.914	0.134	0.090	318±50
1.2	3.574	0.117	0.063	355±79
2.2	5.485	0.115	0.080	304±53
3.2	5.224	0.146	0.093	364±55
4.3	5.173	0.290	0.082	301±51
4.4	5.123	0.181	0.089	349±55
Sample GZ-3				
1.1	2.714	0.057	0.029	228±108
2.1	12.550	1.416	0.207	271±18
3.1	10.732	0.980	0.179	289±23
4.1	2.583	0.048	0.038	307±114
5.1	3.243	0.061	0.042	272±91
6.1	3.298	0.066	0.045	285±90
7.1	3.556	0.036	0.038	234±86
Sample GZ-15				
1.1	6.377	0.123	0.075	249±46
2.1	9.166	0.664	0.145	287±28
3.1	9.203	0.683	0.133	262±28
4.1	10.145	1.299	0.170	266±22
5.1	17.049	1.462	0.273	281±14
6.1	9.613	0.421	0.124	254±29
Session 3-00				
Standard F-5				
1.1	1.102	0.806	0.060	363±84
1.2	5.880	0.953	0.151	375±35
1.3	7.113	1.122	0.158	329±29
1.4	1.911	0.551	0.065	392±85
1.5	6.774	1.084	0.162	352±30
1.6	5.998	1.138	0.153	353±32
1.7	5.767	0.876	0.113	295±37
1.8	6.251	1.103	0.142	323±32
1.9	0.815	0.342	0.033	382±163
Sample ZT-11				
1.1	2.387	0.128	0.044	352±112
1.2	2.350	0.129	0.046	368±113
2.1	3.582	0.405	0.070	322±64
2.2	3.616	0.410	0.069	312±63
3.1	3.582	0.200	0.068	359±74
3.2	3.516	0.205	0.060	322±75
4.1	3.423	0.180	0.065	365±78
4.2	3.369	0.184	0.067	376±79
1.3	0.509	0.280	0.023	364±222
3.3	3.424	0.194	0.064	352±77
3.4	3.456	0.196	0.071	385±76
Sample T-33				
2.1	4.969	0.290	0.082	310±53
2.2	4.290	0.209	0.061	274±63
2.3	4.976	0.284	0.085	324±53
2.4	4.357	0.175	0.049	224±64
4.1	3.520	0.175	0.047	259±77
4.2	3.571	0.157	0.053	290±77
2.5	4.131	0.173	0.054	256±67
4.3	3.799	0.117	0.059	315±75
2.6	4.169	0.188	0.063	297±66
3.1	5.197	0.150	0.060	237±55
3.2	5.256	0.158	0.068	262±54
3.3	4.386	0.179	0.246	1,079±61[a]
3.4	4.315	0.186	0.246	1,090±62[a]
3.5	5.335	0.196	0.155	575±52[a]
3.6	4.406	0.141	0.050	233±65
3.7	4.421	0.125	0.056	262±65
Sample PH-9				
1.1	4.060	0.165	0.072	352±68
2.1	4.149	0.165	0.073	349±67
2.2	4.111	0.174	0.081	389±67
2.3	4.049	0.161	0.059	290±69
2.4	4.069	0.147	0.071	351±69
3.1	2.466	0.761	0.089	405±63
3.2	2.458	0.798	0.081	360±62
1.2	3.568	0.169	0.063	340±76
1.3	3.615	0.168	0.064	343±75
4.1	3.466	0.164	0.071	398±78
4.2	3.495	0.161	0.058	323±78
Sample KS-1				
1.1	5.612	0.898	0.092	243±37
2.1	6.633	0.666	0.111	283±36
2.2	6.733	0.676	0.109	274±35
2.3	6.714	0.700	0.106	265±35
3.1	5.990	0.787	0.100	262±37
Session 6-00				
Standard F-5				
1.1	0.827	0.933	0.055	320±82
1.2	5.795	1.117	0.137	326±33
1.3	6.073	1.090	0.148	345±33
1.4	5.959	1.134	0.141	328±33
1.5	6.194	1.118	0.151	343±32
1.6	5.897	1.139	0.153	358±33
1.7	5.609	1.031	0.145	363±35
1.8	0.812	0.247	0.036	493±193
Sample ZK-25				
1.1	2.595	0.108	0.043	323±106
1.2	2.690	0.103	0.053	392±103
2.1	1.346	0.194	0.022	251±160
2.2	1.357	0.200	0.033	371±156
3.1	2.940	0.147	0.051	331±92
3.2	2.936	0.157	0.045	295±91
4.1	3.108	0.155	0.063	389±87
4.2	3.171	0.150	0.057	350±86
5.1	3.109	0.139	0.045	283±88
5.2	3.110	0.116	0.041	263±90
Sample ZK-3/89				
1.1	3.543	0.238	0.071	368±73
1.2	3.657	0.245	0.067	339±70
2.1	4.755	1.002	0.129	360±39
2.2	4.625	0.790	0.117	365±44
3.1	6.912	0.411	0.146	397±38
3.2	6.944	0.357	0.120	331±39
Sample ID				
1.1	4.479	0.123	0.055	253±64
2.1	4.680	0.152	0.062	270±61
1.2	4.513	0.144	0.059	266±63
2.2	4.866	0.156	0.064	269±58
3.1	3.239	0.149	0.042	252±84
Sample KRO-2				
1.1	2.966	0.217	0.054	329±85
1.2	3.017	0.232	0.050	299±83
2.1	2.272	0.156	0.046	373±113
2.2	2.198	0.143	0.054	451±117
3.1	2.247	0.184	0.036	287±111
3.2	2.223	0.158	0.050	404±114
3.3	2.674	0.242	0.053	344±91
4.1	4.950	0.441	0.112	393±49
5.1	2.264	0.235	0.048	355±104
5.2	2.303	0.239	0.059	428±101
Session 11-01				
Standard F-5				
1.1	9.692	0.754	0.178	329±24
1.2	0.736	0.282	0.024	327±177
1.3	7.482	0.916	0.157	337±28
1.4	7.742	0.767	0.155	339±28
1.5	8.360	0.959	0.180	350±25
1.6	8.668	0.850	0.172	338±25
Sample VF-639				
1.1	4.858	0.097	0.080	344±56
3.2	5.490	0.226	0.097	348±47
1.3	3.914	0.070	0.064	346±70
2.1	3.222	0.071	0.052	339±84

Table 3 (continued)

Grain number/ analysis point	Th	U	Pb	Age	Grain number/ analysis point	Th	U	Pb	Age	Grain number/ analysis point	Th	U	Pb	Age
3.1	4.511	0.195	0.084	364±57	2.3	6.989	0.996	0.164	358±28	3.1	3.854	0.188	0.057	285±65
3.3	5.497	0.227	0.092	330±47	3.3	5.742	0.615	0.128	371±38	4.1	3.644	0.223	0.061	314±67
1.2	4.011	0.089	0.073	379±68	2.3	4.964	2.114	0.178	338±25	2.2	4.046	0.652	0.098	357±47
Sample VF-308					3.3	4.733	0.816	0.110	333±39	3.2	3.801	0.186	0.070	355±66
2.1	3.952	0.165	0.064	321±65	4.3	5.358	0.603	0.117	358±40	1.2	3.929	0.279	0.075	349±60
3.1	3.803	0.104	0.059	318±70	1.1	3.951	1.377	0.138	367±35	5.1	3.950	0.699	0.108	389±47
4.1	4.165	0.291	0.078	340±57	2.1	3.693	3.987	0.236	319±18	6.1	3.520	0.272	0.066	336±66
3.3	5.013	0.093	0.077	325±55	3.1	4.478	0.719	0.104	341±43	5.2	3.854	0.335	0.074	335±59
5.3	3.533	0.055	0.057	343±78	4.1	5.456	0.785	0.133	372±36	**Sample VG-100**				
4.2	4.294	0.577	0.094	342±47	4.2	3.277	3.732	0.236	344±19	1.1	3.901	0.746	0.106	373±46
Sample VF-700					**Sample VF-385**					1.3	3.908	0.342	0.084	372±58
1.2	3.856	1.568	0.133	334±33	1.1	3.465	0.096	0.061	359±77	1.2	3.184	0.532	0.079	359±59
2.3	6.989	0.996	0.164	358±28	2.1	4.156	0.764	0.104	350±44					

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